Volcano monitoring by microgravity and energy budget analysis

Citation for published version:
Brown, GC, Rymer, H & Stevenson, D 1991, 'Volcano monitoring by microgravity and energy budget analysis', *Journal of the Geological Society*, vol. 148, no. 3, pp. 585-593.
https://doi.org/10.1144/gsjgs.148.3.0585

Digital Object Identifier (DOI):
10.1144/gsjgs.148.3.0585

Link:
Link to publication record in Edinburgh Research Explorer

Document Version:
Publisher's PDF, also known as Version of record

Published In:
Journal of the Geological Society

Journal of the Geological Society, London, Vol. **148,** 1991, pp. 585–593, 5 figs, 2 tables. Printed in Northern Ireland

Volcano monitoring by microgravity and energy budget analysis

G. C. BROWN, H. RYMER & D. STEVENSON
Department of Earth Sciences, The Open University, Walton Hall, Milton Keynes, MK7 6AA, UK

Abstract: Microgravity monitoring of active volcanoes can provide evidence of sub-surface mass and/or density changes that precede eruptions. If, in addition, precursory increases in thermal emissions are observed, an integrated mechanistic model for volcanic activity may be developed with the potential for forecasting eruptions. Crater-lake volcanoes provide an interesting target for such studies since thermal output can be monitored simply through lake water calorimetry. Here we summarize 10 years of gravity and thermal data from Poás volcano, Costa Rica. Mass/energy balance calculations demonstrate that, in the steady-state, the large thermal inertia of the crater lake acts as a buffer to short-term changes in the energy input from the cooling magma feeder pipe. Since February 1986, it is postulated that there has been gradual emplacement of a shallow magma intrusion associated with vesiculation and gas loss to the surface. This follows from unambiguous, gravity increases, constrained by elevation control, that are coincident in time with a period of long-term increased energy input to the crater lake. Progressive reduction of the lake volume by evaporation/seepage culminated in an (April 1989) ash eruption providing a good documented record of combined gravity and thermodynamic precursors to volcanic activity.

Active volcanoes are monitored for two main purposes, firstly to improve the understanding of volcano dynamics and secondly to assess volcanic hazard. The latter depends critically on the former and so, as we embark on the International Decade for Natural Hazard Reduction (National Academy Press 1987), it is highly appropriate that the list of conventional monitoring techniques is expanded to include new methods of monitoring the sub-surface flow of mass and energy. This paper concerns microgravity and thermal energy budget monitoring. These are often time-consuming activities, the latter requiring analysis of all the mass and energy flows involved. Ultimately, they may provide a great deal more information of both academic and humanitarian value than, say, studies of surface deformation and temperatures alone. Moreover, intensive monitoring of selected type-volcanoes may generate models with broad applicability, so that the expense of risk evaluation at other targets may be reduced.

Methodology

Microgravity

During the 1980s substantial progress was made in the development of microgravity methods for studying active volcanoes (see reviews by Eggers 1987; Rymer & Brown 1986). Repeated surveys, most notably at Krafla (Iceland), Pacaya (Guatemala), Kilauea (Hawaii) and Rabaul (Papua New Guinea) have demonstrated the potential of microgravity monitoring for interpreting changes in sub-surface mass distribution associated with eruptions. To maximize their value, surveys that are frequent relative to the timescale of natural events are required, with observations at the highest instrumental precision. This can be achieved through a combination of careful field practice coupled with data redundancy based on multiple-instrument and multiple-loop surveys (Rymer 1989). Concurrent elevation control is widely recognised by the authors and others (Tilling 1989) as a crucial factor if interpretation ambiguity is to be reduced and a better understanding of magma dynamics derived. The principle of the method is that, during surface deformation, the change in gravity (Δg, μGal) and elevation (Δh, m) at a series of stations showing different amounts of change (e.g. at varying distances from a centre of activity), will be correlated negatively according to the Bouguer-corrected free-air gradient

$$\Delta g/\Delta h = -(0.3086 - 0.04191\rho) \times 10^3\ \mu\mathrm{Gal\,m^{-1}} \qquad (1)$$

where the density, ρ, is in $\mathrm{Mg\,m^{-3}}$ (see also Fig. 1). This correlation has been observed during the recent eruptive phase at Krafla (Johnsen *et al.* 1980) and, perhaps more interestingly, during the 1982–84 inflation across the Phlegrean Fields (Berrino *et al.* 1984) where no eruption occurred. It follows from equation (1) that any change in sub-surface density will cause a departure from the Bouguer gradient, and this may be indicated by a $\Delta g/\Delta h$ correlation outside the normal range of -200 to $-216\ \mu\mathrm{Gal\,m^{-1}}$ (for $\rho = 2.6 - 2.2\ \mathrm{Mg\,m^{-3}}$). Similarly, any correlation departing from the free-air gradient ($-308.6\ \mu\mathrm{Gal\,m^{-1}}$) implies a change in sub-surface mass. Thus even a Bouguer-corrected free-air gradient correlation must involve the addition or subtraction of mass (during inflation or deflation) in order to maintain a constant overall density; this is because a volume change is implied if an elevation change occurs (see also Fig. 1 caption). It follows that addition of material within the density range of the country rock occurred during inflation in the examples quoted above. Observations at other volcanoes, however, have generated different $\Delta g/\Delta h$ correlations and some of the interpretations of the inferred mass and density changes are indicated in Fig. 1.

Of particular relevance to real-time monitoring are significant changes in gravity with little surface deformation (large Δg, small Δh) observed in 1979–80 at Pacaya, associated with eruptions (Eggers 1983), and in 1983–85 at Poás where no eruption occurred but gravity oscillated about a long-term mean (Rymer & Brown 1984, 1987). These observations were thought to be related to fluctuations in the equilibrium content of exsolved gas within magma bodies beneath the volcano summits, changes that almost certainly were thermally induced. It follows that

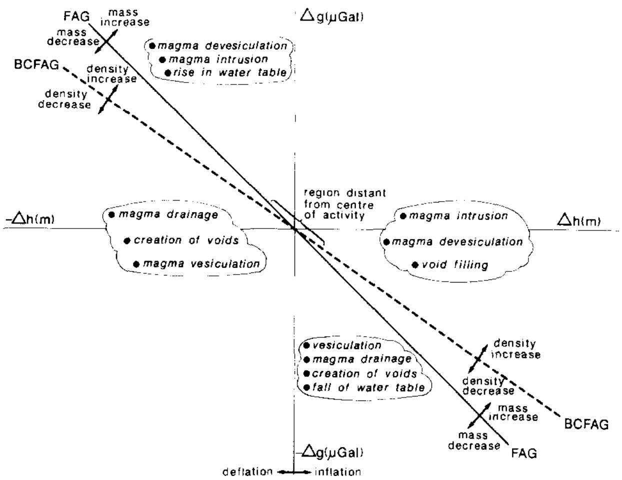

Fig. 1. Normally, gravity and elevation changes are inversely correlated. If no mass change is associated with the elevation change, then gravity will follow the free air gradient (or FAG). Similarly, if no sub-surface density change occurs with elevation changes, then gravity will follow the Bouguer-corrected free air gradient (BCFAG, see Equation 1). The numerical value of the BCFAG depends on density and the FAG is normally taken to be $-308.6\ \mu\text{Gal m}^{-1}$. When observations reveal departures from these 'predicted' gradients, sub-surface mass and density changes are implied. Interpretations in the volcanological context for departures that have been observed in active volcanic regions are shown within the area of the diagram that they fall.

if, instead of simple vesiculation/devesiculation cycles, a potentially catastrophic build-up of magmatic gas pressure or the emplacement of denser magma occurred beneath such a volcano summit this should be detectable at an early stage through a unidirectional change in gravity. Thus microgravity monitoring has the potential both to enrich our understanding of eruption dynamics and to assist with hazard warning. Further theoretical aspects of $\Delta g/\Delta h$ variations are discussed by Savage (1984), while studies of their application to volcanoes, where deformation is rarely simple, are in progress.

Energy budget

Thermal energy budget monitoring of volcanoes is less widely practiced, perhaps because most attention in the literature has been focussed on the thermal and kinetic output of short-duration catastrophic events which reach 10^{18}–10^{19} J (Wilson *et al.* 1978). Many volcanoes maintain long-term efficient cooling systems that produce 10^{8}–10^{10} W continuously; this heat flux may take the form of radiation from a lava lake (e.g. Glaze *et al.* 1989), of gas discharges from fumaroles and boiling mud pools and/or evaporation from hot crater water lakes. While such volcanoes are not immune from more energetic short-term events, their steady-state energy loss over tens or hundreds of years is numerically equivalent to and may replace that of a globally-significant explosive eruption.

More important in the context of monitoring is the development of techniques to measure fluctuations in the continuous thermal output which may herald a catastrophic event. Simple calorimetry provides a good approach to this problem; for example, simply by recording the volumes and chemistry of water in periodic glacier bursts around Vatnajokull (SE Iceland) the steady-state power output of the sub-glacial Grimsvotn volcano has been estimated at 4700–4900 MW (Bjornsson *et al.* 1982).

Volcanoes with hot, acid crater lakes (Table 1) are particularly suitable for calorimetric monitoring, and the incentive is that they are frequently located in populated areas where laharic flows and acid rain are hazardous. We have postulated (Brown *et al.* 1989) that such crater lakes persist in a state of dynamic equilibrium because water losses through evaporation and seepage are balanced by inputs from rainfall and steam inflow from beneath.

Such a lake is the surface expression of a hydrothermal cooling system which transfers heat energy upwards by the convective flow of water/steam from the boundary layer

Table 1. *A comparison of some active crater lakes which have existed for at least 5 years*

Volcano	Surface area of lake (m^2)	Approx. volume of lake (m^3)	Surface temp. (°C)	Comments
Ruapehu, New Zealand	2×10^5	6×10^6	9–60	Post-1970 values, before which lake was deeper and temperature variable
Soufriere, St Vincent	3×10^6 (1970) 2×10^6 (1978)	7.5×10^7 2.5×10^7	22 30	An intra-lake lava eruption in 1971–72 reduced lake size and increased thermal energy output.
El Chichón, Mexico	1.4×10^5	5×10^6	*c.* 25–58	Values for 1982–83; continued cooling since towards ambient temperatures.
Askja, Iceland	2×10^4	2×10^5	20–30	Small explosion crater after 1875 eruption; slow cooling during 1985–88.
Rincón de la Vieja, Costa Rica	3×10^4	c. 1×10^6; depth uncertain	*c.* 40	44° measured in April 1989; since March 1983, this lake has been the site of phreatic eruptions approx. every 12 months, including extensive lahars in 1984.
Poás, Costa Rica	8.4×10^4 (1985) 5×10^4 (3/88) 2.8×10^{4}* (12/88)	1.9×10^6 6×10^5 2×10^5	45 60 82	The lake has gradually been shrinking since 1986.

Note the restricted size range which may suggest that (i) much smaller lakes are unstable because of evaporation while (ii) any thermal flux into larger lakes may be missed because the effects are so diluted.

* Lake area became much more variable as the volume decreased in 1988.

Data sources are Hurst & Dibble (1981) for Ruapehu, Shepherd & Sigurdsson (1982) for Soufriere, Casadevall *et al.* (1984) for El Chichon and our unpublished data for the remainder.

directly above a magma body (Lister 1983). Steam, which may or may not be superheated, discharges as fumaroles into the crater lake where it condenses and heats the lake. Energy is lost at the surface by evaporation, conduction and radiation (Ryan *et al.* 1974). Thus the position of the magma-rock interface is determined by the availability of water and the efficiency of the cooling system relative to the magmatic heat input. The volume of a lake is usually adequate to sustain short-term energy input fluctuations and so it acts as a 'buffer' to inhibit surface activity. Long-term increases in energy input that could precede a catastrophic eruption may well be indicated at an early stage by sustained rises in lake temperature and hence in evaporation rate.

These ideas are supported by the results of thermal energy budget analysis at Poás, described below. Similar thermal monitoring of the crater lake at Ruapehu, New Zealand (Hurst & Dibble 1981), indicates that pre-eruptive periods tend to coincide with low energy input to the lake (<150 MW compared with a normal average of *c.* 400 MW). This phenomenon has led Hurst and co-workers (pers. comm.) to suggest that eruptions at Ruapehu are caused by a blockage in the vent which prevents the free movement of volcanic gases into the lake, a blockage which becomes explosively re-opened. Observations by Shepherd & Sigurdsson (1978, 1982) show that increased temperatures and evaporation rates from the crater lake at Soufriére, St Vincent were clearly linked to the penetration of magma close to the base of the lake, including 1971–72 lava eruptions into the lake. During renewed activity in 1979 vulcanian explosions were caused by magma-water interactions and, once the lake had been removed, quiet lava extrusion ensued.

These few examples of crater lake energy budget monitoring illustrate the immense complexity and uniqueness of each crater lake system. Each volcano to be monitored must therefore be characterized, both in terms of its historical activity and its current dynamic state, before risk assessment by monitoring can be developed.

Monitoring of Poás volcano, Costa Rica

Poás is a composite stratovolcano rising 1400 m above its surroundings within the Cordillera Central in Costa Rica. The last major activity was in 1952–53, when phreatic activity resulted in loss of the previous crater lake and culminated in the growth of a 45 m high pyroclastic cone. The crater lake reappeared in 1967 and apart from fluctuations of a few metres in its level and brief periods of geyser activity (1978 and 1979), the system appeared stable until June 1987 when geysering recommenced. Continuous reduction in the crater lake volume followed until, by April 1989, the lake had virtually disappeared and a 2 km high ash and dry steam column developed over the volcano and was observed intermittently for nearly one month. Extensive static gravity and geological surveys have revealed that the summit area is characterized by at least two debris filled caldera-like structures intruded by the summit crater magma feeder pipe that is the current source of interest (Thorpe *et al.* 1981; Rymer & Brown 1986; Brown *et al.* 1987; Prosser & Carr 1987).

Gravity model

Microgravity monitoring across the summit area during 1983–1985 illustrated the dynamic behaviour of the sub-crater feeder pipe (Rymer & Brown 1984, 1987). The degree of vesiculation within partially molten magma below *c.* 500 m depth was inferred to vary periodically. It became apparent that these variations might absorb large and sudden increased energy inputs from the lower parts of the volcano and therefore contribute to the buffering effect, described earlier, that constrains surface activity. In this way, convectively-transmitted heat pulses from the deeper magma chamber would be dissipated through bubble formation and growth in the magma feeder pipe as well as by changes in the hydrothermal system and lake above.

Throughout our period of observations since 1979 the lake appeared to be in dynamic equilibrium or steady-state (although the rather calm lake surface may have concealed any changes below) until 1985 when gravity at crater bottom stations began to increase with respect to a reference station on the southern flank. Moreover, the crater lake level began to fall consistently until, by 1988, it had dropped by 35 m. Between 1985 and 1989, (Table 2) gravity increased over most of the crater bottom at about 50 μGal a^{-1} and small relative elevation decreases (maximum 30 cm at station G1) were recorded between 1987 and 1989. The residual gravity increases from some typical stations (Fig. 2) show that changes were far greater than can be accounted for by elevation changes alone. After correction for the Bouguer-corrected free-air gradient effect, all the gravity data should plot along the horizontal axis (Δg residual = 0) provided that there are no sub-surface mass or density changes. On the scale of Fig. 2, it would make little difference whether the Bouguer-corrected free-air gradient (for densities of 2.0–2.7 Mg m^{-3}) or the free-air gradient were used for correction. It is clear that variations on the crater rim (R2A) are small, lying close to the limits of resolution (worst case standard errors on the gravity data are 25–40 μGal and for elevation data are 2 cm). However, crater bottom stations such as E3 and D1 (Fig. 2) show statistically significant gravity increases and elevation decreases which reach a maximum at stations on and to the south of the 1953 pyroclastic cone (E6 and E1; Fig. 2 and Table 2). Smaller gravity increases occurred at other crater bottom and rim stations, but gravity *decreases* were observed at station D3 on the north crater bottom. This decrease probably reflects the drop in lake level (removal of water yields: *c.* 50 μGal on the edge of the lake area) and indicates that station D3 is far enough away from the cause of the gravity increases to be unaffected. This makes increases at other stations even more significant since they override the negative effect of the drop in lake level. This is believed to indicate that the effect causing the gravity increases is located towards the southern end of the lake, including the zone beneath the pyroclastic cone.

It is clear from Fig. 2 that there is a large residual net gravity increase above the Bouguer-corrected gradient at crater bottom stations (e.g. 150 μGal at E6, 100 μGal at E1 between 1987 and 1989). The aerial extent and magnitude of the residual gravity increase is used to deduce that an overall mass increase between 10^8 and 10^9 kg occurred beneath the pyroclastic cone area between 1985 and 1989 (Rymer & Brown 1989).

Such an increase in mass could result from the introduction of new magma, perhaps a dendritic intrusion passing upwards through the poorly consolidated and hydrothermally-altered pyroclastic debris characterizing the summit region (Fig. 3). One alternative possibility, that

Table 2. *Microgravity (μGal) and elevation (mm) differences at Poás crater stations relative to values recorded in March 1987 using Lacoste and Romberg instruments G513, G105 and a Nikon DTM-1 theodolite-EDM*

Station no.		March 1985	Feb. 1986	May 1986	March 1987	Jan. 1988	March 1988	Dec. 1988	March 1989
D3	G513	33	—	65	0	−43	−35	—	−29
	G105	—	—	—	0	−60	−114	−126	—
	Elev.	—	—	—	0	−34	−34	−34	−3
E5	G513	—	−12	−12	0	—	163	124	—
	G105	—	—	−36	0	—	185	177	—
	Elev.	—	—	—	0	—	−127	—	−137
E6	G513	−37	26	28	0	72	102	177	239
	G105	—	—	—	0	78	96	151	138
	Elev.	—	—	—	0	—	−130	−110	−112
G1	G513	−145	−86	−94	0	41	68	155	213
	G105	—	—	—	0	94	133	173	198
	Elev.	—	—	—	0	—	−336	—	−306
E1	G513	−77	−66	−39	0	60	72	111	115
	G105	—	—	—	0	51	98	104	118
	Elev.	11	—	—	0	−53	−70	−72	−66
E3	G513	18	80	68	0	54	101	119	186
	G105	—	—	77	0	101	92	95	117
	Elev.	—	—	—	0	−18	−58	−57	−8
D1	G513	15	−86	−96	0	63	97	123	54
	G105	—	—	−31	0	81	92	100	59
	Elev.	—	—	—	0	−34	−47	−19	−26
R2A	G513	11	−29	23	0	39	24	−33	28
	G105	—	—	—	0	43	27	−30	35
	Elev.	—	—	—	0	−15	−35	—	−14
R7	G513	43	−37	—	0	109	76	—	59
	G105	—	—	—	0	94	30	—	66
	Elev.	—	—	—	0	—	−64	—	−14

Each gravity value represents the mean of several difference measurements (over 3–10 days) between the relevant station and a reference point located 1 km from the crater on the southern-flank of Poás. Each elevation value is expressed as a difference from a benchmark at the Mirador overlooking the crater (see Fig. 2).

liquid water now occupies a large volume of porous, pyroclastic pipe-filling debris beneath the crater, where previously there was steam, is considered less likely on two counts. First there was an overall increase in temperature and activity in the adjacent lake area since 1985, which would tend to evaporate more water rather than to condense steam. Secondly, gravity increases over a 10^4 m^2 area require an improbably great 'thickness', 100–1000 m of 10% porous material beneath this area, now to contain liquid rather than gas whereas only 4–40 m of new magma would be required. Nevertheless, in that case it is surprising that no significant inflation occurred at the surface. This suggests a mechanism whereby the new magmatic material must be introduced at the expense of pre-existing less dense material which may be either solid (low density pyroclastics) and/or gaseous (magmatic or hydrothermal vapours). Mechanisms involving the upwards stoping of the chamber roof, of in-situ melting and collapse of low density material and of devesiculation by upwards loss of magmatic gas from the system have some appeal. A second interesting alternative is that pore-filling precipitation from hydrothermal fluids increased the shallow mass without causing inflation. The problem here is that such mass increases might reasonably be expected to have occurred continuously in the upflow region, particularly during cooling rather than heating cycles as recently observed. Neverthess, the possibility of this type of mass change will be kept under review.

In summary, the gravity–height data suggest that by 1986 new magma may have been emplaced some 200–500 m beneath the southern lake area of the active Poás crater (Rymer & Brown 1989). Since gravity continued to increase throughout the 1986–89 period it seems that new mass was added throughout this time interval. If this is envisaged as a new intrusion (cf. Fig. 3) parts of its outer surface may have cooled, fractured and released magmatic gases that were partly responsible for the April 1989 eruptive phase. The gravity increases seen prior to this eruption provide the first documented evidence of precursory effects using this technique. Interestingly, the major changes observed seem to reflect magma emplacement rather than the development of large, potentially-explosive, subsurface gas pockets which we had expected from the earlier evidence of vesiculation cycles (Rymer & Brown 1984, 1987). In contrast, it may be that at least part of the mass increase reflects devesiculation of magma, perhaps by greater leakage of magmatic gases to the surface and their replacement by new magmatic mass.

Thermal model

When the volcano is in a thermodynamic steady-state, the energy input to the hydrothermal system from the cooling magma chamber is balanced by the energy lost (mostly at the summit) at the crater lake and fumaroles. A general thermal model for volcanoes with crater lakes would ideally comprise mass and energy balance considerations for the

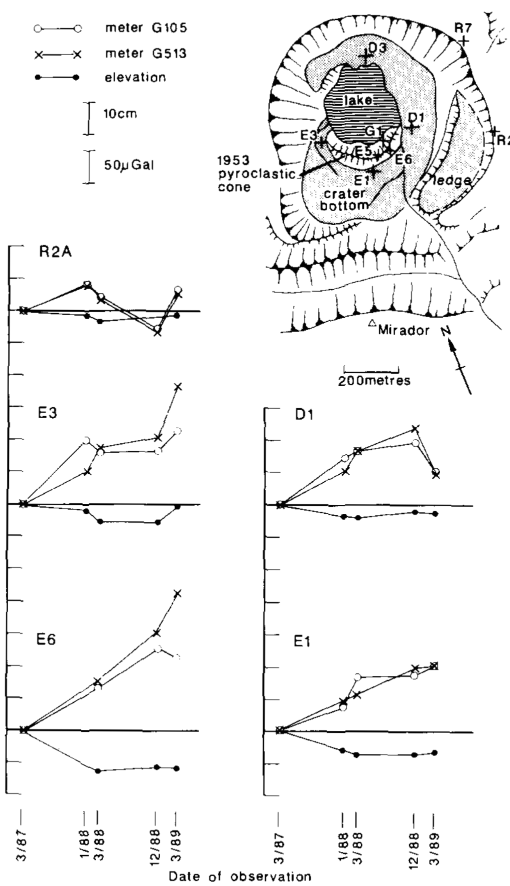

Fig. 2. Elevation and gravity changes for stations at Poás volcano (located on inset) for which most data are available (see Table 2). Elevation data are taken directly from Table 2 but gravity data are *residual* values. Data for each instrument have been corrected for elevation changes using the Bouguer-corrected free-air gradient (Equation 1) and the residual gravity change plotted. Clearly if there were no sub-surface density change, then there would be no residual gravity change and the gravity data would plot along the horizontal axis. No error bars have been drawn since the size of the symbols is greater than or equal to the uncertainty in their position (Rymer & Brown 1989). In the case of the residual gravity data, errors due to an unfortunate choice of density for the BCFAG are insignificant compared with the symbol size.

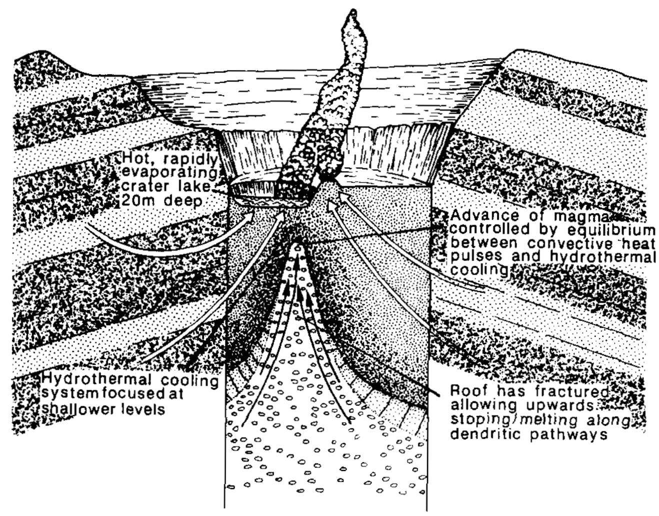

Fig. 3. Schematic cross-section through the summit of Poás volcano as envisaged in 1987, by which time it is proposed that the original boundary layer at depth had fractured extensively allowing melt injection to a new, shallower level where further progress was halted by the increasingly effective hydrothermal cooling system. (Adapted from Rymer & Brown 1989).

entire hydrothermal system. In some cases this involves hydrothermal outflow zones on the volcano flanks while in others most of the energy output seems to be focused at the summit. The part which we describe here is the crater lake (Fig. 4). In the pseudo-steady-state, the lake will fluctuate in temperature and depth, accommodating minor changes in mass and energy inputs and outputs. The fundamental assumption is that the lake is heated from below by the condensation of added steam (probably largely meteoric) which has removed heat from the magma feeder pipe at depth. The total mass inputs to the lake are therefore steam from below and rainwater from above. Mass outputs are seepage through the lake walls and floor and evaporative losses from the lake surface. Any mass change in the system is clearly the sum of these factors (equation 1, Fig. 4). Similarly, energy changes are computed from the difference between inputs from steam and rainwater and outputs due to surface losses by evaporation, conduction and radiation and, less importantly, due to seepage (equation 2, Fig. 4).

Lake level at Poás fell dramatically between 1985 and 1988 and in the process the previous shape of the lake bottom, a complex truncated cone, was revealed. Combining monthly lake temperature and average wind speed data (which, for example, control evaporation rates) with estimates of the lake volume and surface area allows us to deduce the mass and energy balance on a monthly basis (see below). Analysis of the various mechanisms of power output at Poás summit (Brown *et al.* 1989) suggests that, in recent times, summit heat flow, geyser activity and the fumarolic bank (adjacent to the lake on the pyroclastic cone; temperatures in Fig. 5a) are probably relatively minor, and that evaporative losses from the lake surface are the most significant energy sink (50–90%) followed by conductive and radiative losses.

In order to quantify M_{rain} (Fig. 4) raingauge data from the visitors centre c. 1 km south of the active crater, recorded by the Instituto Costarricense de Electricidad through the National Parks rangers, were combined with the size of the catchment area of the entire crater area (*c.* 8×10^5 m^2). Interestingly, during 1988, when the crater lake level fell so dramatically, the highest total rainfall for several years was recorded. By assuming the incoming rain to be at *c.* 15 °C, its enthalpy H_{rain} is estimated (relative to 0 °C) and the total energy ($M_{rain} \times H_{rain}$) deduced from the estimated mass for each month from 1978 to 1989. The energy losses at the lake surface (due to evaporation, conduction and radiation) are estimated using the relationships given in the caption to Fig. 4 using data on lake temperature (Fig. 5b) and surface area. The other variables involved in these equations (windspeed, ambient temperature and humidity) were monitored over an 8 week period to obtain realistic estimates. This leaves 3 unknowns in the two equations for mass and energy balance (Fig. 4) which are the mass and enthalpy of the incoming steam (M_{steam} and H_{steam}) and the mass of the water seeping out of the lake (M_{seep}).

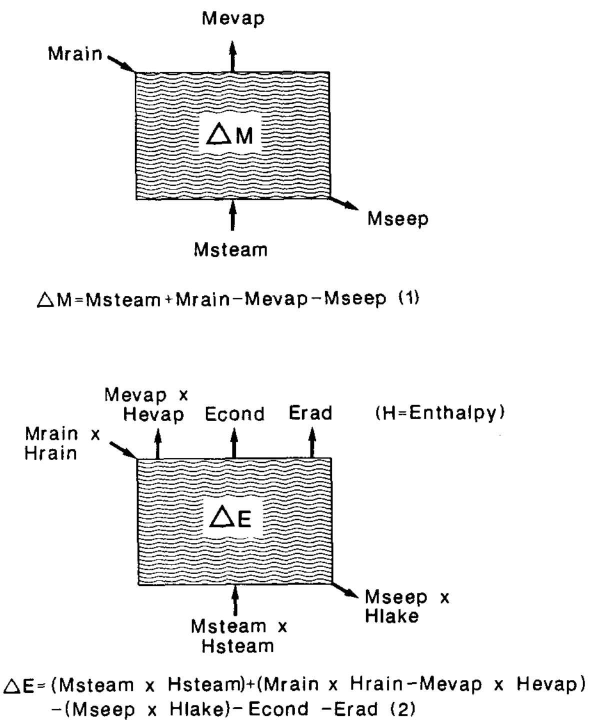

Fig. 4. Schematic illustration of the mass and energy balances used to quantify parameters associated with the crater lake system at Poás volcano. In Equation (1) (mass balance) the rainfall input is calculated from a knowledge of the catchment area combined with local monthly rainfall data, while the mass loss by evaporation is calculated as in (ii) below; this leaves the steam input and seepage output as unknowns. In Equation (2) (energy balance) the rainfall input is as above converted to energy (average temperature 15 °C whereas the enthalpies used in the calculation are relative to 0 °C, for simplicity). The surface energy losses from the lake are as calculated in (i) to (iii) below, leaving the steam input mass and enthalpy and the seepage mass output as unknowns (lake temperature, hence enthalpy is known on a monthly basis). A steam enthalpy was assumed to derive the steam mass input and the seepage mass output using the two equations in the Figure.

Surface energy losses from the lake are estimated as follows:

(i) Net radiative loss

$$E_{rad} = ES(T_s^4 - T_a^4) \cdot A \cdot \Delta t$$

E is emmissivity *c.* 0.9,
S is Stefan–Boltzmann constant $= 5.67 \times 10^{-8}\ \mathrm{Wm^{-2}\ K^{-4}}$,
T_s is absolute surface temperature,
T_a is absolute ambient air temperature *c.* $\approx$288 K,
A is lake's surface area ($\mathrm{m^2}$),
Δt is the time interval (s)

(ii) Evaporative loss

$$E_{evap} = [\mu(T_{sv} - T_{av})^{1/3} + bW_2](e_s - e_2) \cdot A \cdot \Delta t$$

μ is constant $= 2.7\ \mathrm{Wm^{-2}\ mbar^{-1}\ (K)^{-1/3}}$
T_{sv} is virtual surface temperature
T_{av} is virtual ambient air temperature

where virtual temperature $T_v = T/(1 - 0.378e/P)$
e is water vapour pressure at temperature T
P is atmospheric pressure $\approx$740 mbar at Poás

b is constant $= 3.2\ \mathrm{Wm^{-2}\ mbar^{-1}\ (ms^{-1})^{-1}}$
W_2 is windspeed at 2 m height $\approx 2.5\ \mathrm{ms^{-1}}$ (Poás crater floor)
e_s is saturation water vapour pressure temperature T_s

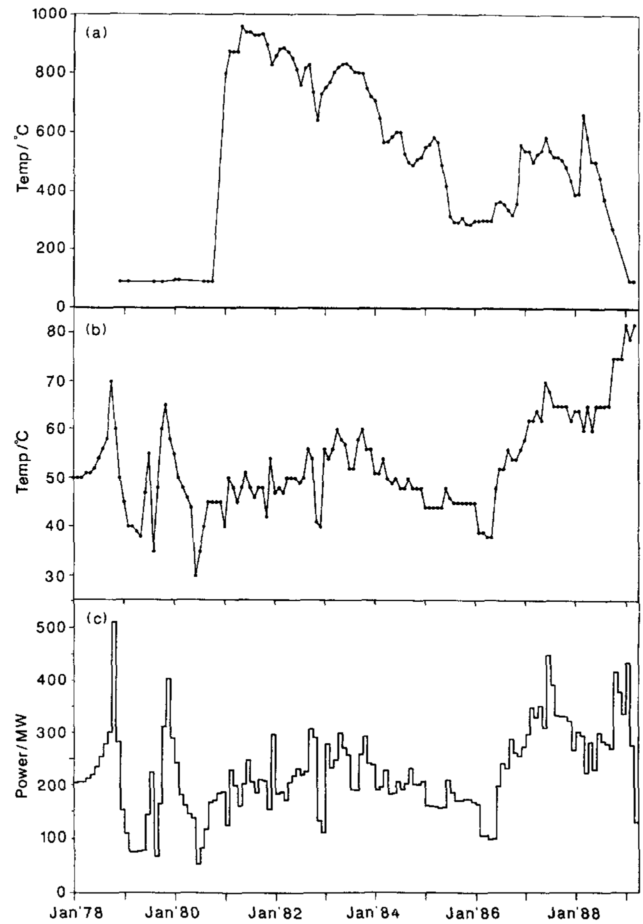

Fig. 5. Monthly plots of (**a**) maximum fumarole temperatures on the pyroclastic cone at Poás, (**b**) lake temperature and (**c**) calculated power input to the lake between 1978 and 1989. Data for (a) and (b) by courtesy of J. Barquero (1988 and pers. comm.); data for (c) result from the model calculations described in Fig. 4.

If we assume the steam enthalpy is that of water boiling at the pressures (4–5 bar) expected at the lake bottom (temperature 150 °C) then an estimate for the steam mass input can be made. The assumption that steam enthalpy corresponds to that on the vapour pressure curve is quite realistic, and provides a maximum estimate of M_{steam}, since numerically there is only a small increase in superheated steam enthalpy above 140 °C. Values of M_{steam} and M_{seep} can then be calculated for each month between January 1978 and March 1989; the former have been converted into energies using an enthalpy of 2.7 MJ $\mathrm{kg^{-1}}$ which gives values

Fig. 4. (*Legend continued*)

e_2 is water vapour pressure 2 m above the lake, arbitrarily say $\approx$20 °C, $\approx$100% humidity, gives $e_2 \approx$23 mbar.

(Mass loss through evaporation is simply $M_{evap} = E_{evap}/H_{evap}$, H_{evap} is the enthalpy of steam at lake temperature $\approx 2.6\ \mathrm{MJkg^{-1}}$)

Conductive loss

(iii) From Bowen's (1926) ratio:

$$E_{cond}/E_{evap} = C[(T_s - T_a)/(e_s - e_a)]$$

C is constant $= 0.61$ mbar $\mathrm{(K)^{-1}}$,
e_a is ambient air vapour pressure.

of $M_{steam} \times H_{steam}$ (Fig. 5c). Significantly, from 1980 to 1986 the average power input from steam appears to have been fairly constant at about 180–200 MW, equivalent to about 70 kg s^{-1}, which agrees well with the steam input value of 80 kg s^{-1} deduced by Brantley *et al.* (1987) on chemical balance arguments.

Prior to the recent period of elevated steam input power, values exceeding 300 MW characterized short periods in late 1978 and 1979 (Fig. 5c). The lake then was about 25–35 m deep, not at its highest level of c. 45 m; these short-term bursts of high steam input are required to balance enhanced surface energy losses caused by the lake temperature rising above 65°C (Fig. 5b). These periods were major deviations from the steady-state steam input power of 200 MW and temperature of *c.* 50 °C; significantly, in 1978 and 1979, geysering activity took place. This was followed by six years of apparently stable activity in the lake area itself, though the power output through the entire crater area fluctuated; for example, fumarole temperatures rose dramatically in late 1980 (Fig. 5a). Particularly heavy rainfall towards the end of the years 1980–82 may have been significant in maintaining the level of the lake, though mass balance calculations (Fig. 5c) also indicate that steam energy input was low during this period perhaps because of enhanced fumarolic activity on the adjacent cone (Fig. 5a). This situation was maintained, but with falling fumarole temperatures, until June 1986 when the steam input to the crater lake began to increase and a level exceeding 100 kg s^{-1} (270 MW) was maintained until 1989. There is no question about the power output at Poás having risen in 1986 as *both* the steam energy input to the lake and the adjacent fumarole temperatures rose at this time (Fig. 5a and c). The average power input to the lake over the period since late 1986 is estimated at 270 MW compared with 180–200 MW during 1980–86.

Interestingly, geysering activity was renewed in mid-1987 when the crater lake temperature again exceeded 65 °C, and the depth had dropped to *c.* 20 m. The reduction in hydrostatic pressure at the lake bed and enhanced lake temperatures may constitute prerequisite conditions for sub-surface steam flash to produce geysering at Poás (cf. similar observations, related to a 60 °C temperature for geysering at the larger Ruapehu crater lake by Hurst & Dibble 1981). The dip in the steam input power graph (Fig. 5c) between mid-1987 and mid-1988 probably is compensated by the additional energy loss directly to the atmosphere associated with geysering (cf. Brown *et al.* 1989). Subsequently the model (Fig. 4) progressively breaks down as direct discharge to the atmosphere from lake bottom fumaroles occurred. It is not surprising, therefore, that by March 1989, the steam input to the lake calculated by the model had fallen to about a quarter of its maximum (1987) value because the lake level had dropped until it comprised only a series of discrete boiling mud pools with fumaroles between. When the lake finally dried up in April 1989, and sediment temperatures rose, sulphur was mobilised to form molten pools on the crater floor because the former lake-bottom sediments reached 116 °C, the melting point of sulphur (Oppenheimer & Stevenson 1989). An eruption of non-juvenile ash (*c.* 500 °C) followed lasting 2–3 weeks and representing a massive power output (probably *greatly* exceeding 100 MW from preliminary kinetic and thermal energy calculations). The subsequent onset of the rainy season (May–December 1989) gradually caused the lake to refill, but only to a rather shallow level (<5 m) and a similar evaporation cycle followed during the early-1990 dry season.

As developed above this model can be used to estimate only the thermal emanations from the crater lake in its steady-state, and one weakness is that mass and energy discharges in the form of fumaroles and occasional geysers at Poás are not included because they are difficult to quantify. An approach to the problem of geyser power (Brown *et al.* 1989) indicates values equivalent to *c.* 15 MW continuous discharge with occasional bursts reaching 400–1000 MW during geysering periods at Poás. Although this leaves unassessed subaerial fumarolic discharges not beneath the lake area, we believe that the lake power input values (i) quantify lake-floor activity well; (ii) give a good lower limit to total volcano power output and, (iii), combined with enhanced fumarole temperatures from 1986 to 1988, suggest that the power output from Poás increased for a sustained period prior to the 1989 eruption.

Discussion of the gravity and thermal models

The first sign of a gravity increase at Poás came in February 1986 when the first data set after March 1985 was obtained; maximum increases of 50–60 μGal occurred at stations E3, E6 and G1 located on and west of the pyroclastic cone (cf. Table 2). The increases continued through 1987–88, though there is some evidence that the rate of increase had peaked by December 1988 and gravity at some stations had begun to fall by March 1989 (see Table 2). Most significantly, however, the period of enhanced power input to the lake (Fig. 5c) corresponds in time with the beginning of the period of gravity and mass increase. The effect on the crater lake level, which characteristically had fluctuated by as much as ±10 m between 1978 and 1985, was that high evaporation/seepage rates between 1987 and 1989 caused a drop in level by 35 m and a long-term rise in temperature (Fig. 5b). Thermal input estimates and lake/fumarole temperature measurements (Fig. 5) show that the energy input to the lake and fumaroles had increased by late 1986 and there is good reason to believe that this new energy level has been sustained ever since. The increases in gravity, however, were progressive and cumulative between 1986 and 1989.

Earlier, the most likely mechanisms of introducing new mass without accompanying inflation were identified as (i) magma emplacement by fracturing, stoping and the downwards displacement of low density, porous pyroclastics (i.e. material backfilling the 1953 vent and hence lying beneath the active crater), (ii) advance of a melting front, collapsing the same material and converting it into magma, and (iii) devesiculation by gas loss from the magma column, with the vacated volume being filled by new magma. Model (i) directly involves an upwards advance of magma through the fractured chamber roof, while in model (ii) the roof zone advances by progressive upwards melting. The size of the intrusion shown in Fig. 3 is 10–100 times larger than required for the new mass (10^8–10^9 kg) alone; and this emphasizes the point that gravity changes can identify only net mass additions and are ambiguous concerning the true size of an intrusion. Thus, if the intrusion were literally the mass calculated from gravity, then just 1–10% of the volume shown would be occupied by dendritic stringers of magma and this represents one extreme variant of models (i) and (ii). At the other extreme, the intrusion could be as large as illustrated provided that the requisite volume of material were displaced downwards (model i) or that a melting front advanced upwards (model ii).

It is instructive to consider the energy content of the new material alone as this will represent a minimum addition of heat to the system shown in Fig. 3. If 10^8–10^9 kg of magma with a latent heat of 3×10^5 J kg^{-1} crystallizes and cools by, say, 500 °C (specific heat capacity 1.1×10^3 J kg^{-1} K^{-1}) it will yield 10^{14}–10^{15} J, or only 1–10 MW over 2–3 years. This is much less heat than required to furnish the enhanced steam energy input between 1986 and 1989 (Fig. 5c), although again, a larger mass of newly molten material than detected gravitationally could be involved. However, the fact that this input to the lake area was sustained until the April 1989 climactic events, and probably since, is evidence against such dramatic cooling of an even larger volume of intruded magma, for, then, convection of heat to the surface would surely become less effective. Indeed, the enhanced power output at Poás over recent years suggests *simply* that the process of heat transport, since 1986, has been *more* effective, either because of more vigorous magmatic convection, hydrothermal convection, or both. If the position of the fractured boundary layer between porous rock and magma is indeed maintained by the balance between heat input from the magma and extraction by the hydrothermal system, then the power increases of 1986 onwards are probably related to upwards migration of magma with a new shallower boundary layer being established (as in Fig. 3). In the discussion that follows it will be assumed that relatively little crystallization occurred between 1986 and 1989.

The argument can be constrained further by recalling that the postulated upwards advance of the boundary layer must have been rapid to satisfy thermal arguments but must also have been accompanied by the *progressive* addition of mass since 1986. A second constraint is that latent heat must be supplied to absorb the stoped blocks (model i) or to advance the melting front (model ii). Thus, on models (i) and (ii) energy will tend to be absorbed, and substantially more heat must be added to the system than the enhanced output observed at the surface. These constraints place severe limitations on the applicability of model (ii) which predicts a progressive rise of energy output, and probably mass, as the melting front advances. However, model (i), (magma injection) can still account for the nature and timing of the observations provided (a) that the previous magma chamber roof cracked in a once-off event, increasing the effectiveness of heat transport, and (b) that new mass has been added progressively as a series of injections through cracks penetrating the sub-crater area, thus helping to sustain the enhanced power output. Model (i) also has the advantage that the latent heat energy required for melting the stoped blocks is not necessarily demanded before the power output is increased at the surface. Nevertheless, ultimately this latent heat must still be provided.

Consideration of these limitations leads to an examination of model (iii). An equally valid explanation of the mass increase follows if the magma chamber roof was fractured and depressurized in 1986, leading to vesiculation, progressive loss of magmatic gas and its replacement by new mass. It is clear from geochemical data (Casertano *et al.* 1985, 1987; Brantley *et al.* 1987) that there is a loss of magmatic gas (principally SO_2, CO_2, and H_2) to the summit at Poás, but dilution by the meteoric system makes this exceedingly difficult to quantify. If, however, a higher magmatic gas component has been lost since 1986, then it might appear that both mass and energy constraints may be satisfied, for the rise of high enthalpy superheated magmatic gases is a particularly effective form of heat transport. Unfortunately, the extra mass of gas lost to create space for the new magmatic mass fails by between 2 and 3 orders of magnitude to account for the extra power output observed. Yet, again, the amount of gas released by a convecting magma body could be much greater than the 10^5–10^7 kg (depending on pressure, density, composition etc.) required to satisfy the mass constraint. (This mass of gas represents a tiny, sub-percent fraction of the likely magma chamber mass beneath Poás). Depressurization would be aided by upwards migration of magma, and so a compromise between models (i) and (iii) is tentatively suggested as the most likely explanation of the temporal mass and energy changes recently observed at Poás.

We propose that the previous boundary layer may well have fractured by 1986 (in fact, late 1985 or early 1986) allowing some combination of (a) magma emplacement by upwards injection and roof stoping to form a dendritic intrusion zone (illustrated by the volume shown in Fig. 3), and (b) enhanced vesiculation and volatile loss from the magma. These processes must have been sustained, perhaps with variable contributions, to account for continued mass increases at least until 1989. It might be envisaged, for example, that initial magma injection was followed by a long period of vesiculation and gas loss, providing the opportunity for further magmatic mass gain. Magma injection led to a shallower boundary layer, thus affording greater access to hot rock by the meteoric system. The consequent increased heat transport would have been aided by the vesiculation process and upwards explusion of magmatic gas. A further consequence of this model is that more effective hydrothermal heat transport needs to be accompanied by a greater heat input from below the boundary layer. Thus the recent event at Poás must ultimately be related to a period of sustained greater heat loss from the magmatic system.

Conclusions

This combined temporal study of gravity, deformation and thermal output from the crater area at Poás volcano has placed various constraints on the trigger mechanisms leading to the April 1989 phreatic eruption columns. Gravity increases (max. *c.* 300 μGal) from 1986 to 1989 have been accompanied by thermal contraction and deflation during cooling of a former fumarolic pyroclastic cone adjacent to the crater lake. The lack of concurrent inflation across the zone of gravity increase is used to establish a mass increase (10^8–10^9 kg) involving the replacement of relatively low density by high density material. The energy flow into the crater lake increased from *c.* 190 to *c.* 270 MW at the beginning of the period of gravity change, and the new, higher level apparently was sustained at least until the April 1989 eruptions. During 1986–89, heat losses mainly by increased evaporation of the hot lake, by geysering and by seepage of hot water from the lake basin lead to the loss of the lake by the end of the 1989 dry season (April–May). Viewed on a short time-scale, loss of the lake's capacity to absorb large changes of heat input provided the immediate trigger for the eruption. But viewed on a longer time-scale, the initial changes of mass and energy in 1985–86 reflected the real causes and, because of the presence of the lake, their effect was delayed by 3 years. The nature of these causes is believed to be connected with the upwards injection of dendritic magma stringers into cracks and fissures, probably accompanied by the downwards displacement of stoped blocks. Vesiculation of the magma throughout the feeder system and subsequent gas loss to the surface probably augmented the transport of additional heat by the hydrothermal system from a shallower magma-solid rock boundary layer established, since 1986, possibly only at c. 200 m depth beneath the lake basin.

Of course, Poás is not a unique volcano in having a crater lake and an active hydrothermal system; it is probable that other similar volcanoes may also have built in buffers that absorb short-term energy fluctuations. As noted earlier, Ruapehu apparently is characterized by low energy output immediately prior to explosive vent clearance (data in Hurst & Dibble 1981). There is a possible parallel at Poás for, after energy flow reached a minimum in 1984–85, the enhancement seen in 1986 is believed to represent a system response by cracking of the magma chamber roof. Clearly, whether this becomes immediately catastrophic will depend on the nature of the system, the efficiency of hydrothermal heat transport, the depth of the magma chamber, the amount of gas liberated and so on. Perhaps a major eruption at Poás would have occurred in 1986 had more energy been released suddenly. This line of reasoning demonstrates that each volcano and each potential eruption is unique in detail, and the volcano must be well characterised before being subject to risk assessment and monitoring. Nevertheless, whatever their final interpretation, the data in this paper demonstrate the value of gravity and thermal monitoring of a volcanic system for an extended period prior to activity: these techniques have considerable value for identifying eruption precursors. It is also clear that monitoring the state of crater lakes at volcanoes like Poás, even when they appear stable, is a valuable pointer to activity because small fluctuations in temperature or level may reflect large changes of energy input.

We thank T. J. O. Sanderson and G. Wadge for constructive reviews; J. Barquero, L. D. Morales and a large number of Earthwatch volunteers for their assistance with data collection. Financial support from Earthwatch, The Centre for Field Research (Watertown, Mass.), the Open University and the UK Natural Environment Research Council (studentship to DS) is gratefully acknowledged.

References

BARQUERO, J. 1988. Changes in fumarole temperature at Volcan Poás. *In*: NEWHALL, C. G. & DZURISIN, D. *Historical Unrest at Large Calderas of the World* vol. 2, USGS Bulletin, **1855**, 878.

BERRINO, G., CORRADO, G., LUONGO, G. & TORO, B. 1984. Ground deformation and gravity changes accompanying the 1982 Pozzuoli uplift. *Bulletin of Volcanology*, **47**, 187–200.

BJORNSSON, H., BJORNSSON, S. & SIGURGEIRSSON, Th. 1982. Penetration of water into hot rock boundaries of magma at Grimsvotn. *Nature*, **295**, 580–591.

BOWEN, I. S. 1926. The ratio of heat losses by conduction and evaporation from air water surface. *Physics Reviews*, **27**, 779–787.

BRANTLEY, S. L., BORGIA, A., ROWE, G., FERNANDEZ, J. F. & REYNOLDS, J. F. 1987. Poás volcano crater lake acts as a condenser for acid metal-rich brine. *Nature*, **330**, 470–472.

BROWN, G. C., RYMER, H. & THORPE, R. S. 1987. Gravity fields and the interpretation of volcanic structure. *Earth and Planetary Science Letters*, **82**, 323–334.

——, ——, DOWDEN, J., KAPADIA, P., STEVENSON, D., BARQUERO, J. & MORALES, L. D. 1989. Energy budget analysis for Poás crater lake: implications for predicting volcanic activity. *Nature*, **339**, 370–373.

CASADEVALL, T. J., de la CRUZ, S., ROSE, W. I., BAGLEY, S., FINNEGAN, D. L. & ZOLLER, W. H. 1984. Crater lake and post eruption hydrothermal activity, El Chichon volcano, Mexico. *Journal of Volcanology and Geothermal Research*, **23**, 169–191.

CASERTANO, L., BORGIA, A., CIGOLINI, C., MORALES, L. D., MONTERO, W., GOMEZ, M. & FERNANDEZ, J. F. 1985. Investigationes Geofisicas y Caracteristicas de las Aguas Hidrotermales: Volcan Poás, Costa Rica. *Geofisica International volcanologia*, **24**, 315–332.

——, ——, ——, ——, —— & FERNANDEZ, J. F. 1987. An integrated dynamic model for the volcanic activity at Poás volcano, Costa Rica. *Bulletin of Volcanology*, **49**, 588–598.

EGGERS, A. A. 1983. Temporal gravity and elevation changes at Pacaya volcano, Guatemala. *Journal of Volcanology and Geothermal Research*, **19**, 223–237.

EGGERS, A. A. 1987. Residual gravity changes and eruption magnitudes. *Journal of Volcanology and Geothermal Research*, **33**, 201–216.

GLAZE, L., FRANCIS, P. W. & ROTHERY, D. A. 1989. Measuring thermal budgets of active volcanoes by satellite remote sensing, *Nature*, **338**, 144–146.

HURST, A. W. & DIBBLE, R. R. 1981. Bathymetry, heat output and convection in Ruapehu crater lake, New Zealand. *Journal of Volcanology and Geothermal Research*, **9**, 215–236.

JOHNSEN, G. V., BJORNSSON, A. & SIGURDSON, S. 1980. Gravity and elevation changes caused by magma movement beneath the Krafla caldera, Northwest Iceland. *Journal of Geophysical Research*, **47**, 132–140.

LISTER, C. R. B. 1983. The basic physics of water penetration into hot rock. *In*: RONA, P. A., BORTROM, A., LAUBIER, L. & SMITH, R. L. (eds) *Hydrothermal Processes at seafloor spreading centres*, Plenum Press 141–167.

NATIONAL ACADEMY PRESS, 1987. *Confronting natural disasters: An international decade of hazard reduction.* Washington, D. C.

OPPENHEIMER, C. & STEVENSON, D. 1989. Liquid sulphur lakes at Poás volcano. *Nature*, **342**, 790–793.

PROSSER, J. T. & CARR, M. J. 1987. Poas volcano, Costa Rica: Geology of the summit region and spatial and temporal variations among the most recent lavas. *Journal of Volcanology and Geothermal Research*, **33**, 131–146.

RYAN, P. J., HALLEMAN, D. R. F. and STOLZENBACH, K. D. 1974. Surface heat loss from cooling ponds. *Water Resources Research*, **10**, 930–938.

RYMER, H. 1989. A contribution to precision microgravity data analysis using LaCoste and Romberg gravity meters. *Geophysical Journal*, **97**, 311–322.

—— & BROWN, G. C. 1984. Periodic gravity changes at Poás volcano, Costa Rica. *Nature*, **311**, 243–245.

—— & —— 1986. Gravity fields and the interpretation of volcanic structures: geological discrimination and temporal evolution. *Journal of Volcanology and Geothermal Research*, **27**, 229–254.

—— & —— 1987. Causes of microgravity change at Poás volcano: An active but non-erupting system. *Bulletin of Volcanology*, **49**, 389–398.

—— & —— 1989. Gravity changes as a precursor to volcanic eruption at Poas volcano, Costa Rica. *Nature*, **342**, 902–905.

SAVAGE, J. C. 1984. Local gravity anomalies produced by dislocation sources. *Journal of Geophysical Research*, **89**, 1945–1952.

SHEPHERD, J. B. & SIGURDSSON, H. 1978. The Sourfriere Crater Lake as a calorimeter *Nature*, **271**, 344–345.

SHEPHERD, J. B. & SIGURDSSON, H. 1982. Mechanism of the 1979 explosive eruption of Soufriere volcano, St. Vincent. *Journal of Volconology and Geothermal Research*, **13**, 119–130.

TILLING, R. I. 1989. Measures of little gravity. *Nature*, **342**, 862–863.

THORPE, R. S., LOCKE, C. A., BROWN, G. C., FRANCIS, P. W. & RANDAL, M. 1981. Magma chamber below Poás volcano, Costa Rica. *Journal of the Geological Society, London*, **138**, 367–373.

WILSON, L., SPARKS, R. S J., HUANG, T. C. and WATKINS, N. D. 1978. The control of volcanic column heights by eruption energetics and dynamics. *Journal of Geophysical Research*, **83**, 1829–1836.

Received 23 January 1990; revised typescript accepted 31 May 1990.